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Forcing a distributed glacier mass balance model with the regional climate model REMO, Part II: Downscaling strategy and results for two Swiss glaciers

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Abstract

Distributed glacier mass balance models are efficient tools for the assessment of climate change impacts on glaciers at regional scales and at high spatial resolution (25-100 m). In general, these models are driven by time series of meteorological parameters that are obtained from a climate station near to a glacier or from climate model output. As most glaciers are located in rugged mountain topography with a high spatial and temporal variability of the meteorological conditions, the challenge is to distribute the point data from a climate station or the grid box values from a regional climate model (RCM) in an appropriate way to the terrain. Here we present an approach that uses normalized grids at the resolution of the mass balance model to capture the spatial variability, and time series from a climate station (Robiei) and an RCM (REMO) to provide a temporal forcing for the mass balance model. The test site near Nufenen Pass (Swiss Alps) covers two glaciers with direct mass balance measurements that are used to demonstrate the approach. The meteorological parameters (temperature, global radiation and precipitation) are obtained for the years 1997 to 1999 (at daily steps) from the climate station Robiei (1898 m a.s.l.) and one grid box of the RCM REMO. The results of the mass balance model agree closely with the measured values and the specific differences in mass balance between the two glaciers and the two balance years are well captured. Despite the disparities in the meteorological forcing from the climate station and REMO, there are only small differences in the modeled mass balances. This gives us confidence that the developed approach of coupling the coarse resolution (18 km) RCM with the high resolution (25 m) mass balance model is suitable and can be applied to other regions as well as to RCM scenario runs.

1. Introduction

Glaciers are key indicators of climate change due to their proximity to melting conditions and the related sensitivity to small climatic fluctuations (Lemke et al., 2007). In particular, the glaciers in the European Alps exhibited strong decreases in extent and thickness during the past two decades, as revealed by satellite observations (Paul et al. 2007a) and by comparison of digital elevation models from different points in time (Bauder et al. 2007; Paul and Haeberli 2008). In principle, there are two closely linked ways of glacier reaction to climate change. Firstly, the mass balance of a glacier is the direct and undelayed response to the annual climatic forcing. Secondly, changes in length represent a delayed, filtered and enhanced response to a more long-term (decadal) climate signal and thus result from a mean mass balance forcing over a longer period of time. Forcing factors for the annual mass balance are all meteorological variables influencing mass and energy exchange with the glacier surface, mainly temperature, global radiation and precipitation (e.g. Greuell and Genthon 2004). Changes in length are delayed as they include aspects of glacier dynamics such as response time, flow velocity and bed topography. These factors are often not well known and could also change over time (e.g. Oerlemans 2001). Thus, changes in length are easy to measure but difficult to interpret, while the opposite applies for mass balance. The reaction of glacier geometry (i.e. extent and surface elevation) to a given (long-term) change in mass balance and thus the climatic forcing, allows us to model or estimate the future length or extent of glaciers according to climate scenarios (e.g. Schmeits and Oerlemans 1997; Raper et al. 2000; Reichert et al. 2001; Paul et al. 2007b; Huss et al. 2008a; Stahl et al. 2008).

As glaciers play an important role at local (hydro-power, tourism), regional (run-off, agriculture) and global scales (sea level rise), a large variety of models has been developed in order

to assess climate change impacts on glaciers and seasonal snow (e.g. Ghan and Shipert 2006). These models are either optimised for a specific glacier (e.g. Schneeberger et al. 2001) and might then perform worse for larger glacier samples, or they are strongly simplified (but still physically robust) and optimised to assess the overall response for several glaciers or large regions, but might fail at the scale of a individual glaciers (e.g. Zemp et al. 2006; Paul et al. 2007b; Rupper and Roe 2008). As a compromise between the two approaches, distributed mass balance models (MBMs) could be applied at the regional scale, given that the governing climatic forcing is known for the entire region and the small scale variability in steep high-mountain topography could be represented by other means.

At the regional scale, the output from RCMs (at a typical resolution of 10-50 km) is relevant for cryospheric impact models (25 m grid spacing). While the meteorological variables as measured at a climate station might lose their correlation with increasing distance to the target site, RCM output provides mean values for an entire region at the respective grid box resolution (Shelly and Hendersson-Sellers 1996). For regional applications this is beneficial and we have thus decided to present the downscaling approach on the example of a distributed MBM. In principle, the approach is also applicable to other impact models (e.g. hydrology, ecology) that require high resolution RCM data in rugged mountain topography.

In this study we use the output of a reanalysis-driven experiment of the RCM REMO (Jacob 2001) for the European Alps, covering the period 1958 to 2002. The experiment is described in detail in Kotlarski et al. (this issue). The selected test site in the Swiss Alps contains two glaciers with available annual mass balance measurements (Gries and Basòdino Glacier) and is mostly covered by one REMO grid box (Fig. 1). Additionally, a high-elevation

climate station (Robiei) is located close to the site. This station is representative for the atmospheric conditions that influence the mass balance of the two selected glaciers. This second part of our study aims to reveal how the differences in the meteorological forcing between REMO and the Robiei climate station influence the modeled glacier mass balance and what kind of corrections are required to use RCM data directly in a high-mountain environment. As such, this study complements part I by analysing a higher spatial and temporal resolution for a specific RCM grid box. For the purpose of this study, we selected two specific and very different mass balance years (1997/98 and 1998/99) rather than a transient long-term forcing. The latter would require to take further glaciological aspects into account (e.g. geometry changes, snow to firn conversion with albedo changes) which would reduce the clarity of the interpretation. Moreover, our focus here is on differences rather than on absolute values. It must also be noted that the RCM output is coupled to the MBM in a one-way mode, i.e. the surface energy balance computed by the MBM does not feedback to the RCM and therefore the MBM acts as a response unit.

In the following, the test site (Section 2) and the principles of the downscaling approach (Section 3) are described. Section 4 introduces the distributed MBM and in Section 5 three meteorological variables from the RCM are compared to the climate station data. The results of the MBM are presented in Section 6 and discussed in Section 7. Major conclusions are provided in the last section.

2. Test site and input data

a. Test site

The test site is located in the south-central part of Switzerland, near the border with Italy in the Nufenenpass region (Fig. 1). It has a size of 27 by 19 km and covers the MeteoSwiss cli-

mate stations Ulrichen (1300 m a.s.l.) and Robiei (1898 m a.s.l.) as well as two glaciers of the Swiss mass balance network, Gries and Basòdino. Gries Glacier is a small valley glacier (size 5.4 km^2 , length 5.7 km) that reaches from 2380 to 3360 m a.s.l. The annual mass balance at this glacier has been measured since 1962 with the direct glaciological method and has been calibrated by aerial photogrammetry about once per decade (Funk et al. 1997). Basòdino Glacier is a small mountain glacier (size 2.7 km^2 , length 1.5 km) that is located about 12 km to the East of Gries Glacier, also has an eastern exposure and is composed of two individual glaciers. The main glacier currently stretches from 2440 to 3220 m a.s.l. and the annual mass balance measurements started in 1992 (WGMS 2007).

Although both glaciers face to the East and are located at a similar altitude, they often experience different mass balances in the same year (WGMS 2007). This might result from slightly different precipitation regimes and the fact that the extent of Basòdino glacier is much closer to a steady-state geometry than the Gries Glacier extent. The region is thus well suited for model validation as these differences in annual mass balance should be reproduced by the MBM and the downscaling approach. It should be mentioned that the pass to the East of Gries Glacier is the lowest point of a major topographic divide, where air masses with North Atlantic origin interact with Mediterranean air. During periods of southerly air flow in autumn, the region around Robiei can receive high amounts of precipitation which are sometimes associated with severe flooding events in the Ticino (Rotunno and Ferretti 2003). For both glaciers a southerly air flow in winter is also the main source of solid precipitation.

The RCM REMO operates on a rotated coordinate system (see Kotlarski et al., this issue). Therefore, prior to our analysis, the RCM output was re-rotated and then converted to geo-

graphic coordinates and projected to the Swiss metric coordinate system (oblique transverse Mercator) using a Geographic Information System (GIS). Digital overlay with the test site revealed that two REMO grid boxes cover the majority of the region (Fig. 1). After some empirical tests with the meteorological data it was decided to use the parameters from box 2797 (running number in the model domain) for the test site. It covers most of the model domain, is located at a similar altitude as the climate station Robiei, and is an inner Alpine grid box which is less influenced by RCM artefacts (see part I of the study). We are aware that such a sharp overlay of RCM boxes is beyond the positional accuracy of the data. However, we are interested in testing the potential of such an approach for impact studies. The other RCM boxes depicted in Fig. 1 are included here as they show interesting differences in the precipitation regime. Averaging over multiple grid boxes has not been performed in this study, although this could be beneficial for certain applications (e.g. Salzmann 2007). The main reason is that any averaging would reduce the temporal variability of the meteorological parameters as simulated by the RCM.

b. Input data

Daily values of temperature (T), global radiation (R) and precipitation (P) are used as regional-scale meteorological forcing factors for the MBM. They are obtained from the Meteoswiss climate station Robiei and from the REMO grid box 2797 for the three-year period 1997-1999 (1095 days). The high-resolution spatial variability of the forcing variables is calculated from a digital elevation model (DEM) with 25 m spatial resolution from Swisstopo (DEM25 in the following) using a lapse rate for T, the radiation code SRAD (Wilson and Galant 2001) for R, and the precipitation climatology from Schwarb et al. (2001) for P.

Potential global radiation (R_{pot}) as calculated with SRAD, uses a standard mid-latitude summer atmospheric profile and accounts for all topographic effects (slope, exposition, shading, sky view factor, etc.) as well as for atmospheric attenuation due to scattering and absorption by aerosols and gases. The radiation as modeled by SRAD is in good agreement with other solar radiation models (Heggem et al. 2001) and direct measurements. In Fig. 2a the distribution of the modeled R_{pot} for Gries Glacier on day 212 (31 July) is illustrated. It can be seen that the glacier receives high amounts of radiation in summer time, as there is little topographic shading and the surface is comparatively flat. Some of the small cirque glaciers to the North are much better protected and most of them do still exist today.

Observation-based climatological precipitation data are digitally available for the entire Alps as mean daily sums per month and per year from Schwarb et al. (2001). It is the most comprehensive and highest-resolution (2 km) data set available to date. Data from more than 6000 precipitation gauges have been used in its creation (Frei and Schär 1998). The applied extrapolation scheme PRISM is described by Daly et al. (1994) and includes, apart from elevation, also slope and aspect facets as predictor variables for precipitation. We assume that the precipitation pattern from the period 1971-1990 is still valid today and resampled the data set by bilinear interpolation to a 25 m cell size grid to avoid sudden precipitation changes from cell to cell and to better account for local trends. The resulting small-scale distribution is of course artificial to some extent, since information on this scale is not contained within the dataset. In Fig. 2b the normalized precipitation grid (all values divided by the mean of the entire region) is shown. The figure reveals that precipitation in the investigated region does not always increase with elevation, but decreases at some sites (e.g. at Basòdino Glacier) and that the local topography exerts a strong influence on the total amount.

3. The downscaling approach

a. Spatio-temporal decomposition

A large number of studies have developed a wide range of statistical and dynamical approaches to downscale climate data to a specific site or point (cf. Wilby and Wigley 1997). If climate station data are used as a meteorological forcing of a distributed MBM, a certain way of extrapolating the measured variables to the entire model domain must be applied in any case. In this regard the forcing has a temporal and a spatial component. We propose here a method of spatio-temporal decomposition. High-resolution (25 m) grids of the most important parameters for glacier mass balance (T, R, P) provide the spatial variability (index xy) in the model domain, while the climate station / the RCM contributes both the regional-scale spatial and the temporal variability (index t). This ensures that the local topographic variability is correctly taken into account while the temporal fluctuations are maintained. The approach assumes stationarity of the topography as well as of the local precipitation gradients and uses the RCM output as a representative mean for the region covered by the RCM grid box.

In order to satisfy basic principles of mass and energy conservation, the spatial downscaling works with normalized values. For precipitation, all values of the 25 m grid from Schwarb et al. (2001) are divided by the mean value of the model domain (yielding P_{xy}), while mean daily global radiation R_t is normalized with the potential global radiation at the respective day and location ($R_{pot,R}$). This allows us to multiply the temporal variability (P_t , R_t) as given by REMO or the climate station Robiei with the spatial variability (P_{xy} , R_{xy}). For temperature, the elevation difference of the respective DEM cell ($Elev_{xy}$) to the elevation of the REMO grid box or the Robiei climate station ($Elev_R$) is calculated and the result is multiplied with the atmospheric lapse rate Γ . This correction is then added to the daily temperature (T_t) from REMO or

Robiei. In a more formal way, the downscaled parameters ($T_{t,xy}$, $R_{t,xy}$, $P_{t,xy}$) are calculated from:

$$T_{t,xy} = T_t + T_{xy} = T_t + \Gamma * (Elev_{xy} - Elev_R) \quad (1)$$

$$R_{t,xy} = (R_t / R_{pot,R}) * R_{xy} \quad (2)$$

$$P_{t,xy} = P_t * P_{t,bias} * P_{xy} \quad (3)$$

where $Elev_{xy}$ = DEM25, $Elev_R$ = elevation of the REMO box or Robiei, R_{xy} = SRAD grids of mean daily potential global radiation, and P_{xy} = the normalized precipitation grid (25 m cells) based on Schwarb et al. (2001). The potential global radiation at the location of the climate station or for the REMO grid box ($R_{pot,R}$) is also calculated with SRAD and extracted for each day of the year from the respective 365 grids. The factor $P_{t,bias}$ is a daily correction factor for the precipitation from REMO that is described in the following.

b. Bias correction of REMO

Apart from the downscaling procedure, it is also necessary to correct for intrinsic RCM biases in the physical representation of atmospheric processes. As shown in part I of this study (Kotlarski et al., this issue), the agreement with climate station data located in the same RCM box could be insufficient at the scale of an individual RCM grid box. However, the comparison also revealed that the monthly variability and mostly also the absolute values are well reproduced by the RCM and that systematic deviations (e.g. radiation in spring at high-elevation sites) are most likely due to unresolved processes in the model physics (e.g. orographic clouds at mountain peaks). Based on these results from part I, we decided to perform a bias correction only for precipitation.

The detected bias in winter temperature (see part I, Fig. 6) was not corrected as it plays a minor role for glacier mass balance. For the same reason, a fixed temperature lapse rate is used to correct for the elevation difference between the RCM box and the DEM25. During summer (when temperature is important) lapse rates are fairly constant (Rolland 2003). The seasonal bias of global radiation in the RCM (see part I, Fig. 12) is also not corrected, because in the current MBM set-up radiation from the RCM is only used in a relative sense, i.e. for calculation of a cloud factor (cf. Section 4b). This helps to maintain the high spatial variability of the modeled radiation and to assess the general influence of the deviations on the modeled mass balance.

For the correction of precipitation we use the observation-based climatology of Schwarb et al. (2001) which refers to the period 1971-1990. Precipitation is calculated by REMO for the same period, and monthly correction factors ($P_{t,bias}$) are obtained for each REMO box by dividing the REMO precipitation by the values from the climatology after spatial aggregation to the size of the respective RCM box (see Section 5). A similar approach of bias correction was proposed by Fröh et al. (2006) for a much larger study region. Due to the high variability of seasonal precipitation sums from RCM box to RCM box (see part I, Fig. 11), the approach of a grid box specific correction factor seems justifiable. We are aware that the bias correction can be interpreted as a model tuning that might lose its validity under future climate conditions. However, the correction with the Schwarb et al. (2001) climatology is at least independent of our meteorologic data set and thus the tuning is not site specific. Moreover, the same approach can be applied to other RCM simulations. As long as precipitation modeled by RCMs has strong biases, a bias correction might be the only possibility of making them applicable for impact modellers.

4. Mass balance modelling

a. Background

A wide range of MBMs with differing complexity have been developed in the past decades (e.g. Arnold et al. 1996; Brock et al. 2000; Klok and Oerlemans 2002; Gerbaux et al. 2005; Hock and Holmgren 2005; Paul et al. 2008). These models are usually designed for specific applications and changing the model set-up (e.g. the time step or spatial resolution) can be quite demanding. The following main differences among the models in use can be identified:

- the spatial resolution of the underlying DEM (e.g. 10-200 m);
- the time step used for the calculation (e.g. hourly, daily);
- the time period modeled (e.g. summer ablation, full mass balance year, several years);
- the complexity of the energy balance computation (involved parameters);
- the number of meteorological variables used for the forcing; and
- the concept for the distribution of point observations to the terrain.

In general, the mass balance of glaciers in a larger region and a specific year is determined by the regional temperature field while the variability from glacier to glacier is dominated by the global radiation receipt and the glacier specific hypsography. Hence, the principle rule for the MBM applied here (cf. Paul et al. 2008) is to consider the spatial variability of the most important factors accurately (vertical temperature change, global radiation, albedo, precipitation) and to parameterize other variables more roughly, for example, by neglecting their limited variability from day-to-day, but considering their much larger variability with altitude (e.g. atmospheric pressure, relative humidity).

When daily time steps are used in the MBM, the model complexity can already be reduced to some degree as certain processes that vary at a sub-daily scale can be neglected. It also increases the representativeness of the measurements at a single point for a larger region when processes with a high temporal fluctuation are considered. For example, cloud cover can show a high variability in space on hourly time scales, but in the course of a day each point in a larger region might have a similar mean cloud cover. The spatial variability of potential global radiation only depends on topography, and can thus be calculated for all cells of a DEM and for each day of the year beforehand using a solar radiation model like SRAD (Wilson and Gallant 2001) or other algorithms (e.g. Corripio 2003). This strongly reduces the computational costs of processing.

The amount of precipitation in mountain regions primarily depends on elevation but varies also on a local scale as topography and the prevailing wind direction exert a strong additional control on the variability (Frei and Schär 1998; Schmidli et al. 2002). The use of a single precipitation gradient for downscaling purposes works well for individual glaciers, but might be less suitable at a regional scale with locally different gradients (Sevruk 1997). When a gridded precipitation climatology at a sufficient spatial resolution is available for such a larger region, it might help to consider the local gradients and thus to improve the calculations in the test site.

b. The distributed mass balance model

The mass balance model (MBM) applied in this study is described in Paul et al. (2008) and has been tested and validated in the same region as here by Machguth et al. (2006b). It is based on the distributed calculation of the energy balance (EB) at the glacier surface according to the formulation by Klok and Oerlemans (2002), but uses simplified approaches for some parameters like turbulent fluxes (which are based on Oerlemans 1991 and 1992) and a constant

0 °C surface temperature. The integration of the MBM starts at the day of the first essential snow falls (around mid-october) with a zero mass balance and zero snow thickness throughout the model domain (cf. Paul et al. 2008). The annual mass balance (MB) is calculated as:

$$MB = \Sigma (\min\{0, -EB/L\} + P_{solid}), \text{ with} \quad (4)$$

$$EB = R (1-\alpha) + LW_{in} - LW_{out} + H_S + H_L \quad (5)$$

In equation (4) the energy balance is positive towards the glacier, L is the latent heat of melt (334 kJ/kg) and P_{solid} is solid precipitation. In equation (5) R is global radiation, α is albedo, LW_{in} and LW_{out} are longwave incoming and outgoing radiation, and H_S and H_L are sensible and latent heat fluxes, respectively. Thus, if a positive energy balance is calculated, the term $-EB/L$ is smaller than 0 and melt of ice or snow occurs in the model. For $T < 1.5$ °C all precipitation falls as snow (e.g. Braun 1991) and thus contributes to P_{solid} . The amount of snow fall for each grid cell is calculated by multiplication of the value from the normalized precipitation grid with the measured / simulated value at Robiei / from the REMO grid box (Eq. 3). Global radiation for each cell of the test site is obtained by multiplying the potential radiation from SRAD for the respective day with a correction factor resulting from a parameterization of the cloud factor (Greuell et al. 1997). This cloud factor is derived by dividing the measured (Robiei) / modeled (REMO) global radiation by the potential global radiation as modeled by SRAD (Eq. 2). For the REMO grid boxes potential radiation was calculated from the REMO topography, i.e. a grid cell of 18 km size at the elevation of box 2797 (2257 m a.s.l.) with an unshaded horizontal surface.

The albedo of the glacier surface without snow has a high spatial variability (due to debris cover) but is temporally nearly constant (Klok and Oerlemans 2004). It is calculated here from a Landsat TM satellite image acquired on 31 Aug 1998 following the approach described by Knapp et al. (1999). At that day the snow extent was at a minimum for the entire period 1980-2002. The bidirectional reflectance distribution function (BRDF) of snow (enhanced forward scattering at low solar elevations) is not considered. For glacier parts that are covered by debris as well as all other terrain a bare rock albedo is calculated using the formulation by Gratton et al. (1993). In the model, the background (bare ice) albedo is unveiled after the seasonal snow has disappeared. The highly variable location of the transient snow line is generated internally with an albedo of freshly fallen snow of 0.75 and an exponential decay in time afterwards (depending on snow thickness and the number of days since the last snowfall).

The longwave outgoing radiation (LW_{out}) is fixed to a 0 °C surface (315.6 W/m^2) for the entire year, neglecting any winter cooling of the snow pack below 0 °C. This results in a slightly too early melt out of the ice as no energy is required to warm the winter snow pack to 0 °C. However, this simplification has little impact on the annual mass balance as Greuell and Oerlemans (1986) have shown. The longwave incoming radiation (LW_{in}) is calculated in the MBM from daily mean temperature and cloud cover with a fixed cloud height following Oerlemans (1991). Turbulent fluxes are calculated using exchange coefficients that do not depend on wind speed but vary with distance from the equilibrium line altitude (ELA) in order to consider an increased surface roughness down-glacier (Oerlemans 1991). The applied ELA value of 2850 m is the mean for both glaciers assuming a zero mass balance. The sensible heat flux only depends on the temperature difference between the air and the glacier surface and the latent heat flux additionally includes water vapour pressure and air pressure, which have both fixed

climatic mean values at sea level (relative humidity 80%, 1013 hPa), but are allowed to vary with elevation and temperature according to Mittaz et al. (2002).

5. Comparison of meteorological parameters

In order to assess whether the three meteorological parameters (T, R and P) as simulated by REMO (cf. Kotlarski et al., this issue) can be used as an input for the mass balance model, a qualitative comparison with the values from the Robiei station at a daily (T, R) and monthly (P) basis is performed. This analysis complements the monthly-based validation in the first part of this study. The coefficients of correlation for T and R are given only for information, as they could emerge from nearly any combination of data points and thus provide little support in the interpretation of the deviations.

a. Temperature

In Fig. 3a the time series of daily mean temperature is shown for the balance year 1997/98 (day 265 to 660). Apart from an obvious shift of the REMO series due to the elevation difference of 360 m, the daily variability is very similar. The related scatter plot of both time series (after a lapse rate correction has been applied to the REMO values) compares individual values and is depicted in Fig. 3b (coefficient of correlation $r=0.93$, linear fit). For temperatures less than 0 °C the deviations are increasingly large, i.e. REMO underestimates temperature in winter time. This finding is consistent with Kotlarski et al. (this issue) who also found a prominent underestimation of winter temperature by REMO. However, this underestimation has only a limited influence on the mass balance of the glacier, as only very few days show large differences and there is in general no snow melt during winter.

b. Radiation

The way in which both of the global radiation time series fluctuate is also similar (Fig. 4a). While large deviations for individual days are found, periods with cloud cover are well reproduced, even during summer. Of course, some degree of divergence from observations has to be expected as REMO is allowed to develop its own meso-scale variability (see below), the data from REMO represent a mean value for a larger region, and finally convective clouds are not resolved by the model but are parameterized. The strong deviations for individual days are also obvious from the scatter plot that is depicted in Fig. 4b ($r=0.8$, linear fit). Apart from that, it seems that measured global radiation is systematically higher by about 20 W/m^2 at many individual days. However, this difference has little influence on the results of the MBM, as daily values from REMO/Robiei are only used in a relative sense, i.e. to calculate a cloud factor. Mean annual cloud factors for the year 1997/98 are 0.29 at Robiei and 0.31 from REMO (0.37 and 0.34 in 1998/99). We have used the global radiation data from REMO without further modification to investigate the influence of such deviations on the modeled mass balance.

c. Precipitation

Especially for precipitation, the comparison of RCM results against observations on a day-to-day basis (compared to the climatological scale) is questionable. In our case, the RCM is forced by re-analysis data only at the lateral boundaries and the atmospheric prognostic variables in the interior of the model domain are not nudged towards the large-scale driving. Therefore, the RCM output does not necessarily match the observations with respect to the location and timing of each individual event, which especially affects daily precipitation sums. For this reason only monthly mean values of daily precipitation are evaluated in Fig. 5a. While the increasing and decreasing monthly trends are reproduced very well in each of the three years, a general underestimation of precipitation by REMO in most of the months is obvious.

This underestimation was also found by Kotlarski et al. (this issue) for many high-elevation regions in the Alps and also for the specific site of Robiei.

In order to find a suitable way of correcting this underestimation, the monthly mean values from the Schwarb et al. (2001) climatology were compared to the REMO values of the same period (1971-1990) within the region covered by each respective RCM grid box (see Section 3b). This independently collected data set of long-term mean values avoids the problems that result when a temporarily or spatially more local tuning is applied. In Fig. 5b the required correction factors are displayed for a selection of six REMO boxes (see Fig. 1). In the case of box 2797 an underestimation of up to a factor of three in winter-time appears. In order to use these corrections on a continuous daily basis, an empirical quadratic function is fitted to the monthly correction factors (yielding $P_{t,bias}$) and applied to the daily precipitation values (P_t) as simulated by REMO. Figure 5b also indicates that REMO boxes 2795, 2796 and 2797 do have a different precipitation regime (seasonally enhanced). This implies that each REMO box might need its own seasonal variable correction factor which confirms that RCM precipitation amounts are not yet practically applicable over large regions in high-mountain terrain using global correction factors.

6. Results of the mass balance model

The mass balance distribution as obtained from the experiment with the Robiei meteo data for Gries and Basòdino Glacier in both balance years is depicted in Figs. 6a-d. The resulting distribution using REMO data as a forcing are nearly identical and are not shown. For better visualization, glacier outlines from 1973 are shown as well. However, mass balance values have been calculated based on satellite-derived glacier extents from 1998. In general, the dif-

ferent conditions in both balance years are well captured and most of the small glaciers at the northern slope have a realistic mass balance pattern as well. Locally, the mass balance distribution varies with potential global radiation (cf. Fig. 2a) and the albedo of the bare glacier ice controls the variability in the ablation region, in particular for Gries Glacier. The satellite derived albedo is probably more realistic than a fixed albedo parameterisation and clearly exhibits the large potential of such albedo maps for calculations on a regional scale (Paul et al. 2008). The more or less complete loss of the accumulation area for both glaciers in the 1997/98 balance year (Figs. 6a and b) as well as the undulating pattern of locally more positive values on Basòdino Glacier could be confirmed by the satellite image (Paul et al. 2005).

A comparison of the individual mass balance values (including a comparison with field data) is provided in Table 1. We do not apply any statistical testing on these values as the sample is too small. For Basòdino Glacier, the mass balance is less negative compared to Gries Glacier in both years, which is in agreement with the observations. However, the differences between both glaciers according to the field data are larger in 1997/98 and smaller in 1998/99 than obtained by the MBM (Table 1). The differences between the mean mass balance as obtained from the forcing with Robiei and REMO data are within acceptable limits (± 0.2 m w.e.) and have opposite directions in both years. This implies that the cumulative effect of the different meteorological forcings on a longer mass balance time-series might be negligible (Machguth et al. 2008). The agreement of modeled and measured values is in general somewhat better for Gries than for Basòdino Glacier. Presumably, the more intense ablation on Gries Glacier is simpler to model compared to Basòdino Glacier, where mass balance might be more strongly influenced by local accumulation processes (e.g. snow redistribution) which are not considered adequately in the model.

In Fig. 7 the mass balance profiles for Gries Glacier (mean values for 50 m elevation bins) as obtained from the modelling and according to field measurements for both years are shown (see Machguth et al. 2006b for curves of Basòdino glacier and a comparison with stake measurements). The curve representing the REMO experiment without correction of the precipitation bias is also shown. This curve clearly reveals a strong overestimation of mass loss, in this case as a consequence of too low precipitation sums (-120%). The mean mass balance for the year 1997/98 with the uncorrected precipitation is about -2.3 m w.e. more negative than with the forcing from the Robiei climate station. This gives a typical mass balance sensitivity of -0.2 m w.e. for a 10% decrease in precipitation. The visible parallel shift of the uncorrected curve in Fig. 7 was also observed in previous studies (e.g. Paul et al. 2008) and illustrates the favour for using precipitation as a tuning factor in mass balance modelling: the mass balance gradient in the ablation area remains nearly unchanged.

The good overall agreement of the modeled and measured curves is quite remarkable, although the latter are regressed by an empirical equation (Funk et al. 1997). Particular strong deviations can be found between 2600 and 2900 m a.s.l. where a larger part of the glacier area is subject to topographic shading (see Fig. 2a), thus reducing the melt in this region. Currently it is not possible to say which profiles are closer to reality, but it is possible that the modeled values are more realistic, because the empirical function smooths the altitudinal variability. However, the MBM might also overestimate mass balances in this region as snow removal due to the frequent small avalanches at steep slopes is not considered in the model. In the upper part of the glacier, the model generates slightly more positive values, which could also be related to an overestimation of precipitation in this region in the Schwarb et al. (2001) data set.

On the other hand, the deviations in the upper part of Basòdino Glacier (not shown) are comparatively small (cf. Fig. 2b).

7. Discussion

The applied downscaling strategy using a normalized grid for P_{xy} and a conversion of the global radiation R_t to a cloud factor worked successfully and is flexible in its application (e.g. Paul et al. 2009). The proposed spatio-temporal decomposition allows to preserve the temporal variability of the meteorological forcing parameters as provided by the RCM while maintaining the high spatial variability of the parameters in rugged mountain topography on a local scale. If we assume that the spatial variability of the used high-resolution observational precipitation data set is robust on decadal time-scales, it can also be used for downscaling under future climate conditions. This would allow to apply the presented method also in long-term simulations without introducing a systematic bias. Such a bias can occur with downscaling approaches that calibrate RCM data to measurements of a reference period that might not reflect future climate conditions. However, other studies found that the precipitation data set by Schwarb et al. (2001) needs to be regionally adjusted as well (Huss et al. 2008b, Machguth et al. 2009). As a prerequisite, the DEM used for interpolation of temperature and calculation of potential global radiation must have a spatial resolution that is appropriate for the size of the considered glaciers and an accuracy that allows a proper calculation of shadowing effects from mountain crests (e.g. Klok and Oerlemans 2002, Arnold et al. 2006). The most critical part of the modelling process is then the availability of accurate, gridded precipitation data, which are at this high spatial resolution and quality only available for the Alps. In other parts of the world a simple precipitation gradient might be successful as well or other data sources may be utilized, such as humidity values from radiosondes (Rasmussen and Conway 2003).

The temperature extrapolation could be enhanced by a seasonally varying lapse rate (e.g., as applied in Kotlarski et al., this issue), but this will only give a minor change of the winter balance, as temperature has in general little influence on glacier mass balance during winter. It appears that global radiation can be taken from REMO without further correction, at least when the values are used to derive a cloud factor. The strong deviations of radiation values observed at individual days (see Fig. 4b) are of minor importance for the modeled mass balance as long as differences are approximately normally distributed. However, the comparison with monthly values from mountain stations which are influenced by local orographic clouds, has shown that not every climate station or REMO grid box can be used to drive the MBM (see part I of the study). The precipitation data from REMO show an important bias, especially in high-elevation regions and a proper correction might be difficult in other parts of the world due to a lack of observation data. However, it might still be possible to obtain good model results when the annual sum and the altitudinal gradient are known, as shown by Oerlemans (1992). The exact timing of individual precipitation events is also of minor importance as long as a natural temporal pattern and correct seasonal amounts are generated by the RCM. Indeed, the applied empirical correction of the total amounts of precipitation based on the monthly climatology by Schwarb et al. (2001) would change for each grid box (Fig. 5b) and other correction concepts might be more appropriate when larger regions (i.e. several RCM grid boxes) are covered (cf. Machguth et al. 2009).

The applied MBM works with several simplifications: it includes only the most important processes in detail and the energy and mass fluxes are not closed, i.e. the sum of all related terms is different from zero. A more sophisticated calculation of key variables (e.g. relative humidity, pressure, cloud height, seasonal lapse rate) - including their temporal variability -

could well enhance the reliability of the model, but the pattern of mass balance distribution would not change in general. This can be explained by the insignificant influence of these variables on mass balance (e.g. Oerlemans 2001; Klok and Oerlemans 2002) as they tend to cancel each other and because their variability with elevation is much larger than with time (e.g. for pressure and temperature). Model intercomparisons also indicate that more complex parameterisations of the energy balance have little influence on the calculated mass balance of individual years (Hock et al. 2007). The MBM used here is open to incorporate such data and the output from RCMs is especially suited for assimilation by such impact models. However, there is a strong need to improve MBMs for processes in the accumulation region (snow redistribution by wind and avalanches) rather than for the well understood glacier melt processes (Greuell and Genthon 2004; Hock 2005). Hence, our modelling approach will likely generate reasonable values for glaciers in the Alps where the annual mass balance is largely driven by summer balance. Deviations for modeled winter balance could be much larger, but currently the required distributed data sets for validating modeled winter balance are only rarely available (e.g. Dadić et al. 2008; Machguth et al. 2006a). Moreover, the errors in the field measurements at individual points might be large (e.g. Escher-Vetter et al. 2009) and thus not appropriate for a sound model validation.

A future step for improvement of the MBM is certainly the discrimination between clean to slightly dirty glacier ice and debris-covered ice (i.e. rocky material protecting the ice from radiation). In principle, this could be implemented by using a map of debris-covered area from an automated classification of glacier ice with multispectral Landsat TM imagery (Paul et al. 2004). Such glacier maps usually include the dirty ice (with increased ablation) but not the densely debris-covered parts (with decreased ablation). The selection of the correct melt reduc-

tion factor (e.g. around 0.5) is then another task, but several studies exist on that topic and could help in finding a suitable range of values (e.g. Nicholson and Benn 2006). Compared to the available measured mass balance profiles, the modeled profiles are more variable and might even be more realistic than the fitted measured curves, but this is difficult to assess when the pattern observed in the field is not known. Considering this, there is little value in tuning the model until results fit better to the measurements as the deviations are within the limits of uncertainty (Machguth et al. 2008). In the future, validation of MBMs might increasingly rely on a comparison of spatial mass balance patterns (Paul et al. 2009).

8. Conclusions

Using the example of a distributed glacier mass balance model we have presented a method for downscaling RCM data in rugged high-mountain terrain. In this approach, the spatial variability of atmospheric parameters is represented by normalized grids (temperature, precipitation) and raster data sets which are calculated beforehand (potential global radiation), while the temporal variability is based on daily time series from the RCM REMO or the climate station Robiei. The latter do generally agree well with data from REMO (for temperature and global radiation), but radiation shows strong deviations on individual days. Total amounts of precipitation are generally underestimated by the RCM and have to be corrected differently for individual RCM boxes. In the present case, an empirical function is used that has been derived from a comparison of the REMO precipitation with long-term monthly mean values from a gridded high-resolution (2 km) observational climatology. This data set also proved to be very useful in a region with a complex precipitation pattern and is at least available for the entire Alps. The data from the Robiei climate station are distributed to the model domain by the same approach, but without the bias correction for precipitation.

Although the MBM is forced with daily means from only three climatic parameters, the modeled mass balances for the two glaciers, the two years and the two different forcings (observations and RCM) are in good agreement with field measurements. The observed deviations of the modeled mass balance profiles may in part be attributed to the regression curve applied to the field measurements which do thus not fully reflect the local variability. The strong deviations of the RCM derived atmospheric variables for individual days seem to have a minor influence on the overall results. This is promising for an application of the approach in other glacierized regions. The application of correction factors from a gridded high-resolution climatology to adjust RCM precipitation is mandatory, but can also be regarded as a suitable way to correct them without using a local tuning. In this sense, cryospheric impact models can help to assess the quality of RCM output in mountain terrain.

The difference in spatial resolution of the RCM (18 km) and the MBM (25 m) is of minor importance for the regional application presented here, as the RCM provides spatially representative mean values of atmospheric variables at a regional scale. From this point-of-view and for this specific application, we therefore conclude that future improvements of RCMs should first focus on a reliable estimation of regional precipitation amounts rather than on increasing the model resolution (though both aspects might be connected to each other). Future work with the downscaling approach presented here will include the application to further regions, other time series and more than one RCM box, as well as to model the impacts of future climate change scenarios on glacier mass balance.

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Figure captions

Fig. 1: Overview of the test site (yellow square) in the Swiss Alps at the border to Italy (cyan) with the numbered boxes of the REMO grid (thick green), the position of the ANETZ climate stations from Meteoswiss (orange) and glacier areas (blue) from the 1973 Swiss glacier inventory. North is at top, the size of the test site is 27 by 19 km. The background shows a Landsat Thematic Mapper satellite image from 31 August 1998, which is also used for calculation of glacier albedo. The satellite image was obtained from NPOC.

Fig. 2: a) Mean daily potential global radiation for day of the year 212 (31 July) as obtained from the radiation model SRAD for Gries Glacier. Glacier outlines (thick) and 200 m elevation contours (thin) are superimposed. b) Precipitation factors as derived from the annual sums of the Schwarb et al. (2001) climatology and resampled bilinearly to 25 m cell size. Glacier outlines (thick) and 200 m elevation contours (thin) are superimposed. A local precipitation maxima/minima is visible near the climate stations Robiei and Ulrichen (yellow dots).

Fig. 3: a) Time series of mean daily temperature in the balance year 1997/98 for REMO grid box 2797 (grey) and the climate station Robiei (black). The elevation difference (360 m) is not corrected in this graph. b) Scatter plot of the temperature data from Fig. 3a including a lapse rate correction for the REMO data.

Fig. 4: a) Time series of mean daily global radiation in the balance year 1997/98 for REMO grid box 2797 (grey) and the climate station Robiei (black). b) As a) but displayed in a scatter plot. The grey line is the identity.

Fig. 5: a) Monthly mean values of modeled and measured precipitation for REMO box 2797 and the climate station Robiei in all three years. b) Monthly correction factors for all six RCM grid boxes (see Fig. 1) as obtained from the Schwarb et al. (2001) climatology (period 1971-1990). The daily factors for REMO box 2797 were obtained by an empirically fitted quadratic function for each day of the year.

Figs. 6: Modeled mass balance distribution for both glaciers (Gries: a and c, Basòdino: b and d) and balance years (1997/98: a and b, 1998/99: c and d) as obtained with the meteorological data from Robiei. Glacier outlines from 1973 are white and 100 m elevation contours are black. The red area at the terminus of Gries Glacier is due to a lake that has expanded since 1973. This area is not considered in the calculation of the mean mass balance.

Fig. 7: Mass balance profiles for Gries Glacier from field measurements and as modeled with the REMO and the Robiei time series. The thin black line is the resulting mass balance profile when the REMO precipitation data are not corrected.

Table

Table 1: Overview of modelled and measured glacier mass balances in m water equivalent. REMO precipitation amounts were corrected prior to forcing the MBM (see text).

	1997/98			1998/99		
	Robiei	REMO corrected	Field	Robiei	REMO corrected	Field
Gries	-1.54	-1.69	-1.66	-0.59	-0.39	-0.58
Basòdino	-1.25	-1.31	-1.07	-0.22	-0.09	-0.44

Figures

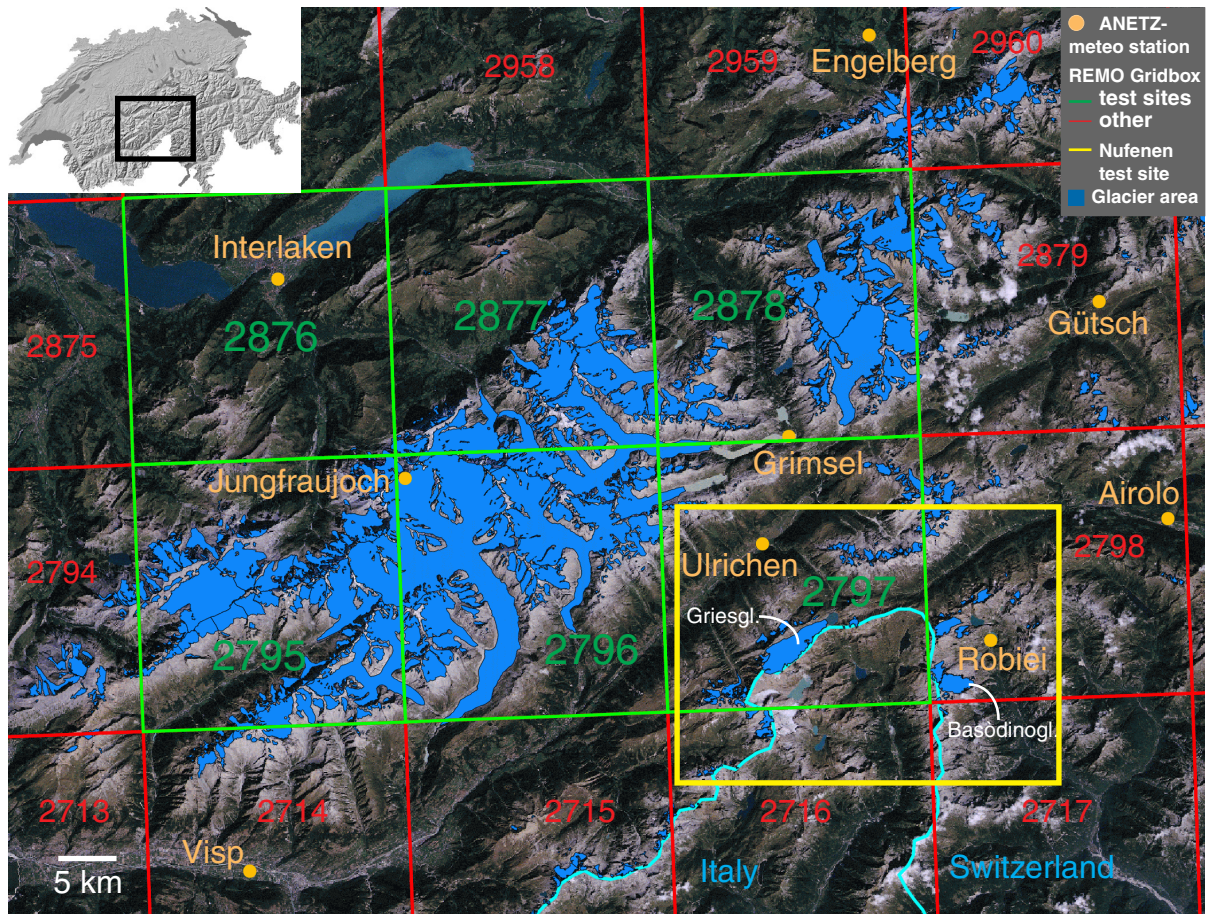


Fig. 1: Overview of the test site (yellow square) in the Swiss Alps at the border to Italy (cyan) with the numbered boxes of the REMO grid (thick green), the position of the ANETZ climate stations from Meteoswiss (orange) and glacier areas (blue) from the 1973 Swiss glacier inventory. North is at top, the size of the test site is 27 by 19 km. The background shows a Landsat Thematic Mapper satellite image from 31 August 1998, which is also used for calculation of glacier albedo. The satellite image was obtained from NPOC.

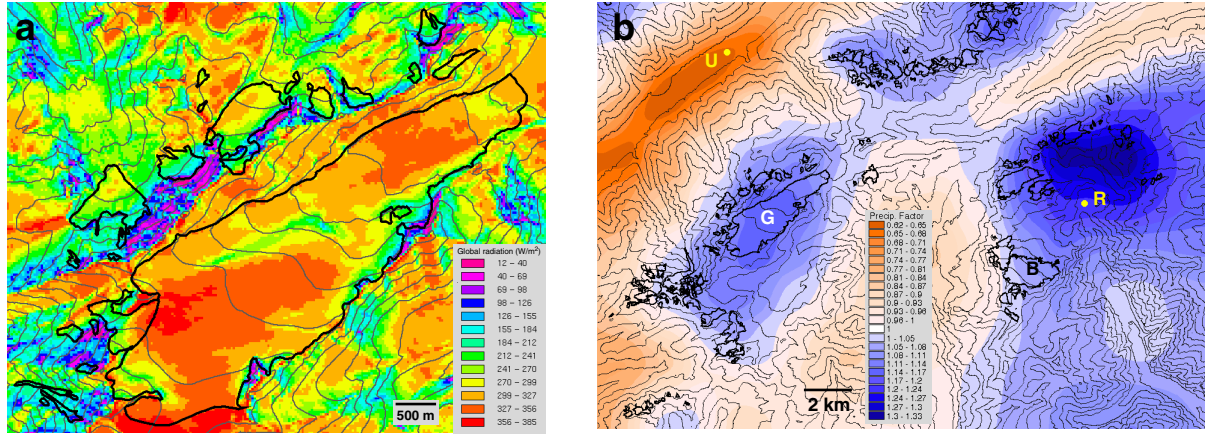


Fig. 2: a) Mean daily potential global radiation for day of the year 212 (31 July) as obtained from the radiation model SRAD for Gries Glacier. Glacier outlines (thick) and 200 m elevation contours (thin) are superimposed. b) Precipitation factors as derived from the annual sums of the Schwarb et al. (2001) climatology and resampled bilinearly to 25 m cell size. Glacier outlines (thick) and 200 m elevation contours (thin) are superimposed. A local precipitation maxima/minima is visible near the climate stations Robiei (R) / Ulrichen (U).

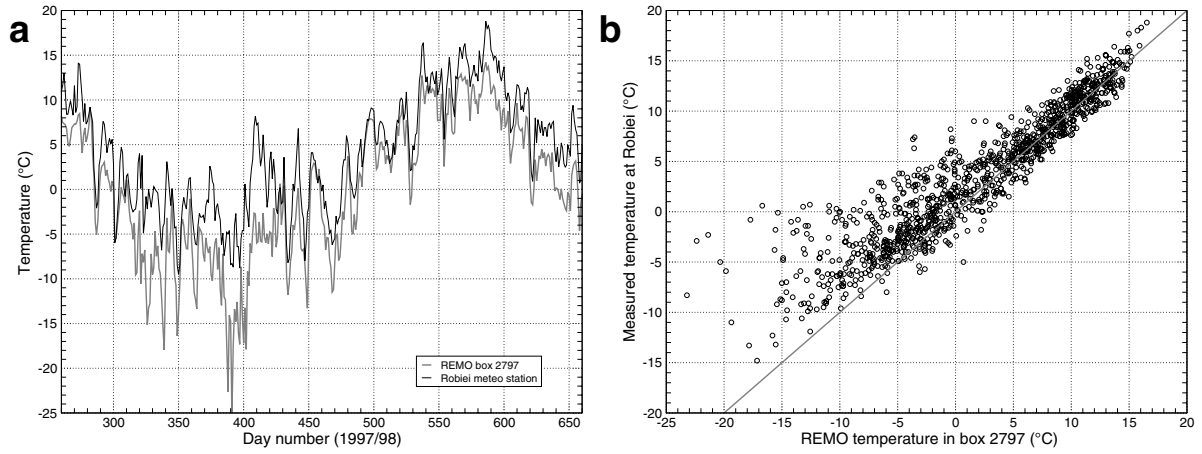


Fig. 3: a) Time series of mean daily temperature in the balance year 1997/98 for REMO grid box 2797 (grey) and the climate station Robiei (black). The elevation difference (360 m) is not corrected in this graph. b) Scatter plot of the temperature data from Fig. 3a including a lapse rate correction for the REMO data.

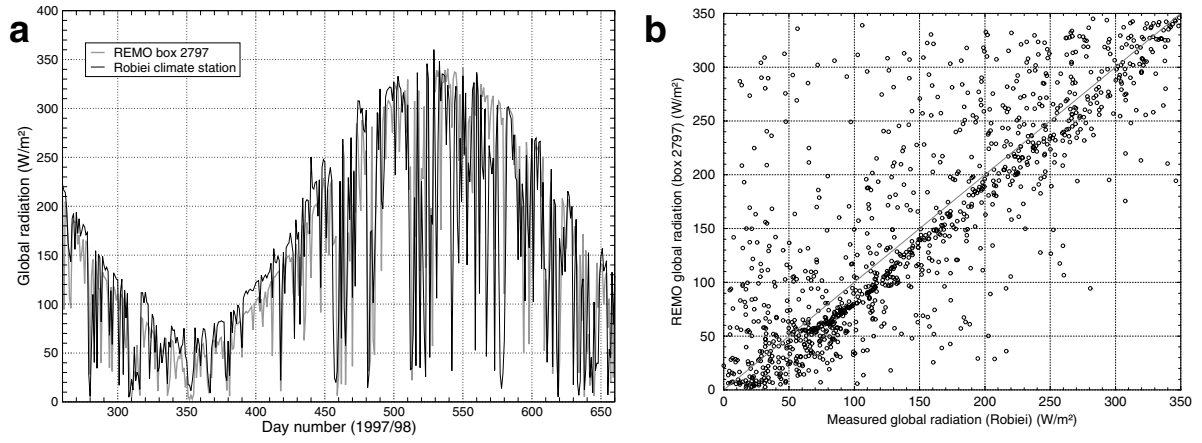


Fig. 4: a) Time series of mean daily global radiation in the balance year 1997/98 for REMO grid box 2797 (grey) and the climate station Robiei (black). b) As a) but displayed in a scatter plot. The grey line is the identity.

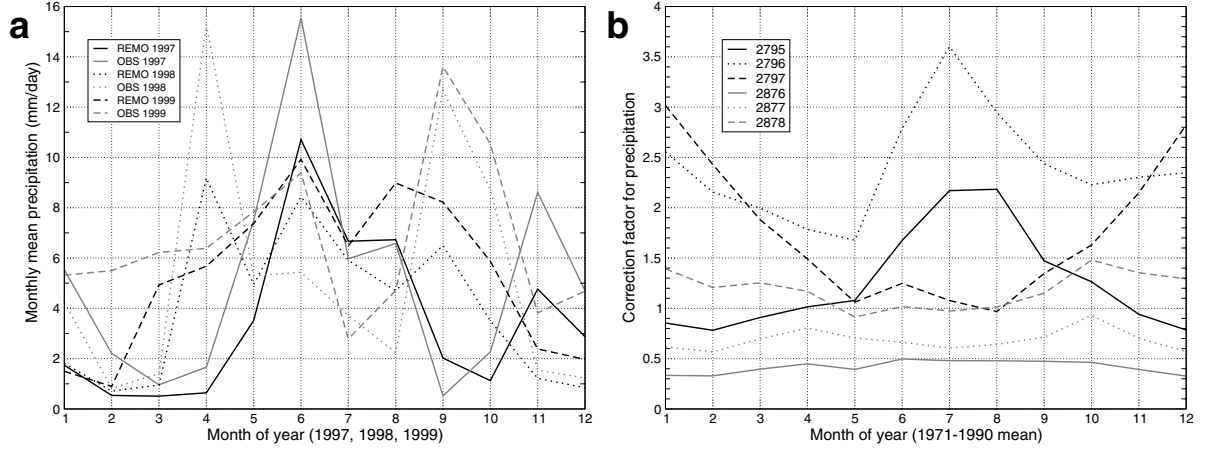
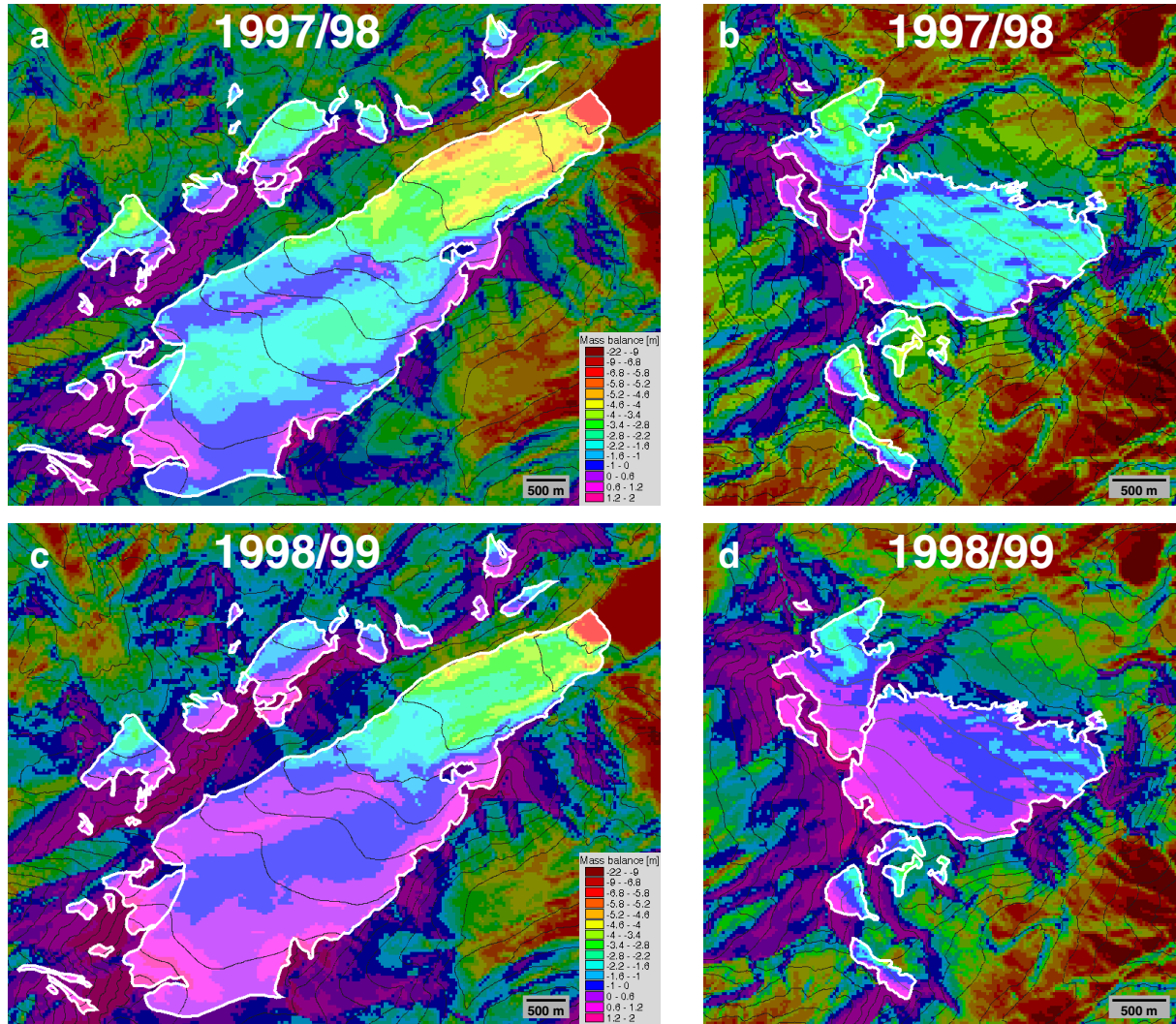


Fig. 5: a) Monthly mean values of modeled and measured precipitation for REMO box 2797 and the climate station Robiei in all three years. b) Monthly correction factors for all six RCM grid boxes (see Fig. 1) as obtained from the Schwarb et al. (2001) climatology (period 1971-1990). The daily factors for REMO box 2797 were obtained by an empirically fitted quadratic function for each day of the year.



Figs. 6: Modeled mass balance distribution for both glaciers (Gries: a and c, Basòdino: b and d) and balance years (1997/98: a and b, 1998/99: c and d) as obtained with the meteorological data from Robiei. Glacier outlines from 1973 are white and 100 m elevation contours are black. The red area at the terminus of Gries Glacier is due to a lake that has expanded since 1973. This area is not considered in the calculation of the mean mass balance.

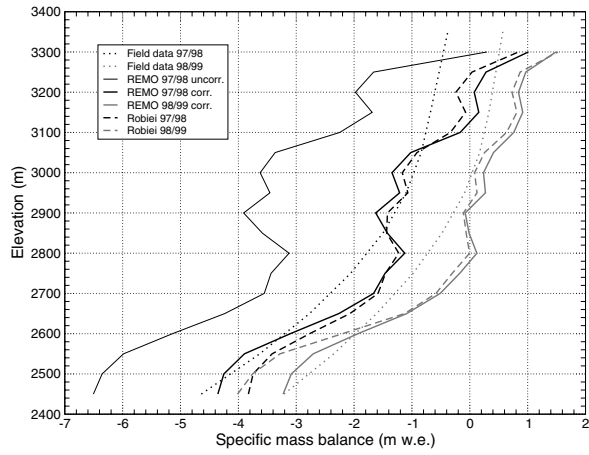


Fig. 7: Mass balance profiles for Gries Glacier from field measurements and as modeled with the REMO and the Robiei time series. The thin black line is the resulting mass balance profile when the REMO precipitation data are not corrected.